Chapter 21

Paleozoic paleosols

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Introduction

Soils and life are so intimately interconnected that it is difficult to conceive of many kinds of life without soils or of many kinds of soils without life. Nevertheless, there is life in both sea and air that is only indirectly dependent on soil. There also are soils on mountain tops and in Antarctic deserts that are only minimally affected by life. The surficially altered surfaces of the Moon, Mars and Venus also have been called soils and, so far, all appear devoid of life. Comparison between these various modern soils and organisms is one way of assessing the interrelationship between soil and life. But there is another way. The fossil record of soils, particularly those of Paleozoic age, some 245 to 570 million years old, can be viewed as a long term natural experiment in soil formation under increasingly complex terrestrial ecosystems. During this period of geological time, multicellular land plants and animals appeared and diversified to create most of the main kinds of ecosystems now gracing the land. Along with them, diversified different kinds of soils. Paleozoic paleosols can be regarded as trace fossils of ancient ecosystems, providing a long term record of increasing biotic influence on soil formation.

The Paleozoic diversification of soils could be assessed by a variety of methods. Most of these are quite premature, given the level of available information on Paleozoic paleosols (Retallack, 1986a) and the progress yet to be made in understanding how much of soil-forming processes are preserved in such ancient and altered paleosols (Retallack and Wright, 1990). For these reasons, it is not yet possible to make compilations of named paleosols, comparable with plots of changing diversity through geological time available for fossil land plants (Niklas et al., 1985), insects (Sepkoski and Hulver, 1985) or vertebrates (Benton, 1985). It is, however, feasible to analyze the first appearance of major kinds of paleosols. These are sufficiently distinctive that they can be recognized among a large number of paleosols examined in the field or assessed from scientific literature. This preliminary effort to document first appearances of paleosol types serves as a reconnaissance outline of

soil diversification during Paleozoic time. Later studies may refine these conclusions and perhaps prove that some peculiar Paleozoic paleosols represent extinct types of soil. Until such a surplus of information accrues, we must settle only for tracing the origins of modern soil types, such as the orders of the U.S. Soil Conservation Service or "Soil Taxonomy". There are only ten of these orders, and only nine need to be considered here because Mollisols are unknown in deposits older than Eocene, some 53 million years ago (Retallack, 1982).

In considering the diversification of life on land during Paleozoic time, the tabulation of numbers of genera or families of terrestrial fossils already mentioned are not readily comparable with soil types. The environmental and historical constraints of terrestrial ecosystems are such that their complement of species, genera and families appears almost random (Bernado and Webb, 1977). Instead, emphasis will be placed here on the overall structure of plant communities; that is to say, on distinctive kinds of plant formations. These are not the same as rock formations in geology, but rather refer to the general appearance and structure of vegetation. Wooded grassland (or savanna in a very restricted sense) can appear quite similar in South America, Africa and Australia, despite the very different plant and animal species involved on each continent (Cole, 1985). It is plant formations that are reflected best in the major features of soils (Foth and Schafer, 1980), rather than plant species or communities in a taxonomic sense. My nomenclature of plant formations follows that developed in recent vegetation mapping of Africa (White, 1983), with a few additions (Retallack, 1990). The Paleozoic was a time of transition from simple, visually unimpressive, microbial earths and rocklands, to many of the plant formations found today, with a concomitant diversification of soils.

Entisols

These kinds of soils showing minimal profile development probably formed well back in geological time. They must have existed as early stages in the development of better developed paleosols now known as old as 3400 million years (Lowe, 1983). Fossilized Entisols are difficult to recognize in Precambrian rocks because their degree of weathering is slight. Fossil root traces and burrows are needed to distinguish ancient Entisols from associated sedimentary, metamorphic or igneous rocks. In a few cases, some as old as 2300 million years, periglacial soil features such as sand wedges can be used to recognize Precambrian Entisols (Cryorthents and Cryopsamments: Williams, 1986).

Among the most ancient Entisols known to have supported multicellular land life are burrowed red sandstones and siltstones, such as the Faust Flat silty clay paleosol (a Fluvent) of Late Ordovician (Ashgillian) age in the Juniata Formation of central Pennsylvania (Figure 21.1). These paleosols show no evidence of gleying, and presumably formed in well-drained parts of an extensive coastal floodplain (Feakes and

Fig. 21.1. Outcrop of type Potters Mills clay (below upper arrow) and type Faust Flat silty clay (below lower arrow) paleosols in the Late Ordovician (Ashgillian) Juniata Formation near Potters Mills, Pennsylvania (at 40°45'32"N and 77°37'04"W on U.S. Geological Survey 7.5 minute Spring Mills Quadrangle). Hammer handle is 25 cm.

Retallack, 1988). The conspicuous (1–21 mm diameter) and abundant burrows are superficially similar to those of millipedes (Retallack and Feakes, 1987). There is no evidence of fossil roots or rhizomes, but there are surficial reduction spots that may have formed around buried organic matter. Any plants present were presumably non-vascular and perhaps liverwort-like (Retallack, 1985). Similar paleosols are found in rocks of most succeeding geological periods (Retallack, 1986a). Their plants probably were early successional stages in the colonization of bare ground.

Preservation of fossil plants in Entisols is much better where soil development and plant decay have been inhibited by waterlogging, by extreme cold or by fire. A very early example of a salt-marsh paleosol has been found in the Early Silurian (Llandoverian) Frenchville Formation near Stockholm, Maine. It contains erect axes

of the early land plant Eohostimella heathana (Schopf et al., 1966; Niklas, 1982). The plant-bearing horizon is associated with marine fossils.

Sandy Entisols (Psamments) have been found among deposits of coastal dunes in the Mid-Pennsylvanian (Desmoinesian, equivalent to Moscovian) Hermosa Formation at several localities in Utah, as well as in the Early Permian (Wolfcampian, or Sakmarian) Cedar Mesa Formation of Utah and Fountain Formation of Wyoming (Loope, 1988). The size distribution of calcareous root traces in these paleosols are indications of desert shrublands and scrub.

Silty Entisols with shallow carbonaceous root traces (Cryofluvents) have been found at several levels in earliest Permian (Asselian and Sakmarian) fluvioglacial deposits in eastern New South Wales, Australia, which was at that time close to the south pole. Some of these weakly developed paleosols, such as those near Lochinvar, contain low-diversity assemblages of leaves of *Botrychiopsis*, which may have formed a kind of polar tundra (Retallack, 1980). Other paleosols at slightly higher stratigraphic levels in this postglacial sequence have a more diverse assemblage of leaves dominated by *Gangamopteris*, in an ancient vegetation that was an early version of taiga. The similar alpine plant formations of fellfield and krummholz also may date back to late Paleozoic time, but no evidence of them has yet been discovered.

Some shallow stony soils (Orthents) support fireprone shrublands, variously known as chaparral (U.S.A.), maquis (Europe) and matorral (Chile). Such soils and vegetation are indicated by dolines in paleokarst of latest Triassic or early Jurassic (Rhaeto-Liassic) age filled with fossil charcoal (Harris, 1957). Fossil charcoal has been found as far back in the rock record as earliest Mississippian (Tournaisian: Cope and Chaloner, 1985), but the association of charcoal with bedrock paleosols has not yet been established so far back in time.

Inceptisols

These soils of somewhat stronger development than Entisols also can be difficult to recognize in Precambrian rocks. Surprisingly, many Precambrian paleosols that are exceptionally thick and well developed often turn out to be best classified as Inceptisols, mainly because they lack the diagnostic properties of other modern soil orders (Retallack, 1986a). As more is learned of Precambrian paleosols, some of these profiles probably will become recognized as extinct orders.

The most ancient Inceptisols with remains of multicellular life yet recorded are of Late Ordovician age in Pennsylvania and Nova Scotia. The Potters Mills clay paleosol in central Pennsylvania is a common kind of profile in fluvial deposits of the Late Ordovician (Ashgillian) Juniata Formation (Figure 21.1). These profiles are distinguished by a surficial network of conspicuous (1–15 mm diameter) burrows and a subsurface horizon of comparably-sized calcrete nodules (Feakes and Retallack, 1988). The nodules truncate and are truncated by the burrow walls indicating

that the burrows are part of the original soil. From their geological setting these paleosols appear to have formed on well drained terraces of an extensive piedmont, flanking a major mountain range of folded sedimentary and metamorphic rocks. The nature of these paleosols, together with information from marine faunas of equivalent age, are evidence of a subhumid, perhaps seasonally dry, tropical climate. As in comparable modern soils, the paleosols show evidence of calcification, ferruginization and clay formation. The fossil burrows in these Ordovician paleosols provide some evidence of large land animals, probably millipede-like arthropods (Retallack and Feakes, 1987). There is no clear evidence of the likely vegetation of these soils, such as traces of roots, rhizomes or rhizoids. If this negative evidence can be used to exclude the possibility of vascular land plants, the paleosols could have supported cyanobacteria and other microbes, or even forms of a comparable grade of evolution to lichens and liverworts. For the following reasons, my own opinion is that they supported thallose multicellular plants. The fecal pellets in burrows and the abundance of the burrows indicate a productive ecosystem that included a considerable biomass of solid food. Furthermore, trace elements normally associated with organic matter in soils (such as Li, Nb, Ni, P, Sr and Y) show a pattern of surficial depletion that is more like that found in soils with a surface cover of plants than those with an interstitial fauna or a richly humified surface layer.

Looking for evidence of this ancient vegetation further afield, land plant spores of a kind comparable to those of living liverworts are widespread in Late Ordovician rocks (Gray, 1985), well before spores of vascular land plants and megafossil remains of land plants appear in geologically younger (Silurian) rocks. There also is another paleosol of Late Ordovician (Caradocian) age developed on an andesite lava flow in Nova Scotia, which has reduction spots in mounds of paleosol between erosion scours filled with redeposited material (Figure 21.2). These features have been interpreted (Dewey in Boucot et al., 1974) as evidence for clumps of early land plants which locally stabilized the soil against erosion, and whose remnant organic matter decayed anaerobically to create the green reduction spots. Lichens and liverworts are common in very dry sites today: on tree trunks, on cliffs and among alpine boulder fields. Extensive cover of such thallose vegetation in well-drained parts of lowland fluvial settings, as envisaged here for Late Ordovician times, is not found today. This distinctive archaic kind of plant formation can be called a polsterland (Retallack, 1990).

The advent of vascular land plants in well-drained soils also may be recorded in Paleozoic paleosols. Late Silurian (Ludlovian) paleosols of the Bloomsburg Formation near Palmerton, Pennsylvania, are in many ways similar to those of Late Ordovician age already described from Pennsylvania. These Silurian paleosols also formed on quartzo-feldspathic floodplains flanking a major mountain range in a subhumid tropical climate. Unlike the Ordovician paleosols, the Silurian ones show irregular surficial bioturbation that is similar to what one would expect for rhizomes of land plants (Retallack, 1985). These traces are indistinct and have not been stud-
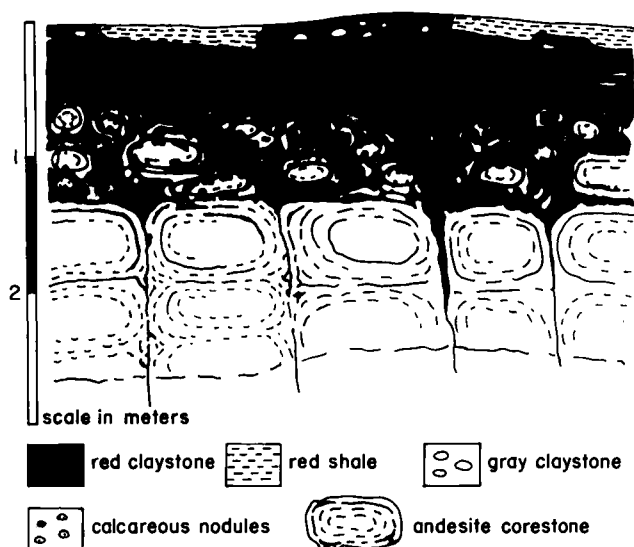
Fig. 21.2. Field sketch of surficial reduction spots and scour-and-fill structure, and of subsurface calcrete nodules and corestones in a Late Ordovician paleosol developed on an andesite flow near Arisaig, Nova Scotia (from Boucot et al., 1974).

ied in such detail or numbers to be certain of this interpretation. Nevertheless, they are a promising new line of evidence in the quest for understanding early vegetation on dry land. Colonization of the land by vascular plants in Silurian time also is in evidence from globally widespread and diverse fossil spores, sporangia and stems of vascular land plants of this geological age (Gray, 1985; Gensel and Andrews, 1984). Such vegetation can be imagined as comparable to fields of bracken fern (*Pteridium aquilinum*) that spring up after forest fires, or to patches of clubmosses (*Lycopodium*) in alpine meadows. This distinctive early kind of vegetation can be called brakeland (Retallack, 1990).

Polsterlands and brakelands of the early Paleozoic were supplanted by more modern kinds of vegetation and soils during mid-Paleozoic time. The large woody roots of trees leave clear and distinctive traces in soils, and a good record of the advent of woodlands and forests can be expected from paleosols. There is paleobotanical evidence for a variety of trees by Middle Devonian (Givetian) time (Banks et al., 1985) and paleosols of this geological age confirm that these trees formed woodlands and forests. The Peas Eddy clay paleosol in the Late Devonian (Frasnian) Walton Formation near Hancock, New York, has been studied in some detail. The large root traces penetrating this purple-brown siltstone have not entirely disrupted relict bedding (Retallack, 1985). Although it was not a well developed paleosol, there is evidence of slight clay illuviation and of incipient translocation of iron: both indicators that it was an Inceptisol. This profile also formed in quartzo-feldspathic

alluvium of a major mountain range, but it is non-calcareous, unlike associated Late Devonian paleosols of former floodplains away from streams. Early forest of this weakly podzolized paleosol appears to have formed galleries along streams.

By Late Devonian time there are indications of considerable local variation in vegetation. In the same Late Devonian (Frasnian) outcrops as the Peas Eddy clay there are other fossilized Inceptisols. These have carbonized woody root traces in a gray massive clay and a horizon of pyrite nodules. Such soils (Sulfaquepts) are typical of intertidal woodland, or mangal. Exactly what kind of trees formed this early mangal has not yet been investigated, but such reducing environments are very favorable for preserving plant fossils. Slightly older (Middle Devonian or Givetian) siltstones in New York containing marine fossils have within them large stumps and rootlets of trees (Johnson, 1972). These stumps of *Eospermatopteris erianium* have been considered progymnospermous, but may in fact be lycopods (Scheckler, pers. commun., 1988). A variety of other kinds of plants formed mangal vegetation long before the golden fern (*Acrostichum aureum*), palm (*Nipa*) and angiosperms (*Avicennia*, and *Rhizophora* among others) that now are recognized as mangroves (DiMichele et al., 1987).

Histosols

Peaty soils of permanently waterlogged ground have an excellent fossil record in the form of coal seams and petrified peat, within which often are preserved fossils of the plants that grew in them. There may be evidence of freshwater marshes as early as Late Silurian (Ludlovian) if local accumulations of early land plants in the Bloomsburg Formation of Pennsylvania (Willard, 1938; Strother and Traverse, 1979) prove to be sufficiently thick and in place of growth. Freshwater marsh certainly existed by Early Devonian (Seigenian) time. The Rhynie Chert of this age in Scotland is a well known permineralized peat preserving in exquisite detail a variety of herbaceous early vascular (*Rhynia, Horneophyton, Asteroxylon*) and non-vascular (*Aglaophyton, Lyonophyton*) plants (Kidston and Lang, 1921; Edwards and Edwards, 1986). Of about the same geological age near Barzass, in central Siberian U.S.S.R., there are thick (4 m) cuticle coals of the early land plant *Orestovia devonica* (Stach et al., 1975; Krassilov, 1981).

The appearance of trees in waterlogged peaty substrates may have occurred somewhat later in geological time. The oldest woody coal of a thickness that would qualify as a Histosol after considering its compaction due to burial is of latest Devonian (Famennian) age in West Virginia (Gillespie et al., 1981). It is supported salt-marsh of the zygopterid fern *Rhacophyton ceratangium*. Other pyritic nodules with remains of tree cordaites in coals of Late Carboniferous age appear to represent ancient mangal vegetation (Raymond and Phillips, 1983). Woody coals are abundant and widespread in Carboniferous, Permian, Cretaceous and early Tertiary rocks (DiMichele et al., 1987).

Most coals appear to have formed in acidic swamps and marshes. However, there are in the United States and western Europe, many Carboniferous coals containing calcite or dolomite nodules that may have formed during accumulation of the peat. Various lines of evidence can be marshalled to support the idea that these were soil nodules rather than clasts eroded from other rocks or phenomena of alteration of the peat after its burial (Scott and Rex, 1985). The most compelling evidence is the exquisite cellular preservation of uncrushed fleshy fossil plant tissues, even at the base of coal seams where plant debris in modern comparably thick peats is compacted. These ancient wooded wetlands were thus chemically neutral to alkaline. Thus, vegetation of coal balls with tree ferns and tree lycopods should be called carr, rather than swamp. Other coal balls containing only small ferns (Rex and Scott, 1987) represent a vegetation most like modern fens. Alkaline wetlands with fen and carr vegetation are widespread today, but only a few show evidence of carbonate permineralization of plant tissues, and in no known case is preservation as exquisite as in Carboniferous coal balls (Retallack, 1986b). Subhumid climate and nearby limestone bedrock promote fens in wetlands, but it also could be that phenolic compounds and other acidifying agents of Paleozoic plants were less effective in controlling soil pH than are those of many modern plants. Certainly the tree lycopods (*Lepidodendron*) and horsetails (*Calamites*), so well represented in Pennsylvanian coal balls, were very different plants from the woody seed ferns and conifers of Permian and geologically younger coals.

Aridisols

Shallow, stony, salty and calcareous soils, like those now forming in deserts, have a long fossil record. The earliest calcrete-bearing paleosols are some 1800 million years old in the Canadian Northwest Territories, and other examples provide a fossil record of these soils reaching to the present (Retallack, 1986a). When these soils began to support different kinds of vegetation is a project amenable to study, considering the common preservation of root traces in modern calcretes (Klappa, 1980). A comprehensive study of this kind has not yet been done. Many of the great array of calcareous paleosols from the latest Silurian (Pridolian) to earliest Carboniferous (Tournaisian) Old Red Sandstone of Britain, probably were Aridisols. Some contain fossil root traces and burrows (Allen, 1986). More detailed studies have been reported of Aridisols from South Wales of Early Carboniferous (Visean) age (Wright, 1987). By this geological time such soils certainly had been invaded by plants of moderate stature. Root traces in the Heatherslade Geosol at Three Cliffs Bay and Miskin reach a diameter of 3 mm (Wright, 1986). Associated with these root traces are needles of calcite of a kind identical to those now formed by fungi associated with roots. Aridisols containing large woody root traces are known in Early Carboniferous (Visean) limestones of Kentucky (Ettensohn et al.,

1988), and Triassic sandstones of Texas (Blodgett, 1988), Connecticut (Hubert, 1977) and West Germany (Martins and Pfefferkorn, 1988). Fossil Aridisols in the latest Jurassic (Tithonian) limestones of Dorset in England, are notable for fossil stumps and root traces as well as plant fossils of a dry conifer-bennettite woodland (Francis, 1986).

Vertisols

Swelling clay soils also are of great antiquity. The oldest example currently known is a paleosol some 2200 million years old in a deep roadcut near the town of Waterval Onder, South Africa (Retallack, 1986c). In this large exposure can be seen the swalelike thickening and thinning of its surface horizon and deep cracking of the ridges (mukkara structure) that is a characteristic subsurface expression of gilgai microrelief. Large exposures are needed to confidently identify fossil Vertisols, so that paleosols more ancient than the Waterval Onder profile but known only from drill cores (Grandstaff et al., 1986) may yet turn out to be fossil Vertisols.

When Vertisols began to support plants remains an open question. Probably this occurred at least by Devonian time. Fossil Vertisols with tubular features are known among Devonian paleosols of Britain (Allen, 1986) and Antarctica (McPherson, 1979). Detailed studies of their roots or burrows remain to be reported. It is conceivable that they supported shrublands, wooded shrublands or dry woodland. Open grasslands and wooded grasslands, so widespread on modern Vertisols, probably were a much later addition to the array of vegetation on Vertisols. Grasses have been found fossilized no older than Cretaceous, and the main kind of grassland soils (Mollisol) are known no older than Eocene (Retallack, 1982).

Alfisols

Base-rich soils with a clayey subsurface (Bt) horizon today are largely under woodland and forest. These soils also seem to date back to about the time of the earliest forests. The earliest example known to me is in the uppermost Old Red Sandstone on the rock platform north of Pease Bay, southeastern Scotland (Figure 21.3). This alluvial sequence is of Late Devonian (Famennian) age, judging from associated fossil fish of *Holoptychius nobilissimus* (Craig, 1975). The paleosol profile has a relict shale bed, much disrupted by burrows and root traces. This clayey part of the profile is not an argillic horizon, but the down-profile increase in clay and clay skins in the sandy overlying layer does appear to be pedogenic. The paleosol also has carbonate nodules, in this case dolomitic and mainly in that part of the profile below the clayey layer. In petrographic thin sections the nodules consist of rhombohedral dolomite (Figure 21.4), a crystal habit often taken as evidence of dolomitization of

Fig. 21.3. The type Pease Bay loamy sand paleosol beneath cross-bedded sandstone in the Late Devonian (Fammennian) Old Red Sandstone on the rock platform north of Pease Bay, Scotland (British National Grid NT802713). Sledge hammer handle on Bt horizon is 23 cm.
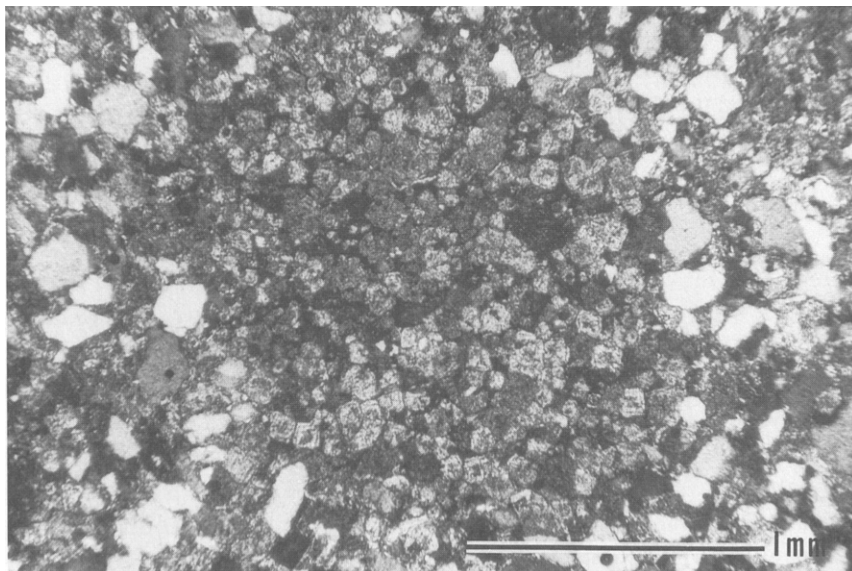

Fig. 21.4. Photomicrograph, under crossed nicols, of recrystallized dolomite (rhombs) in small calcrete nodule of the type Pease Bay loam sand paleosol in the Late Devonian (Famennian) Old Red Sandstone near Pease Bay, Scotland. Scale bar is 1 mm.

other carbonates after burial (Zenger and Mazzullo, 1982). Dolomite is found in very few modern soils (Cline, 1955), but is common in paleosols of Devonian age and older (Retallack, 1986a). The nodules themselves are original features, judging from their avoidance by some of the burrows and root traces. Whether they were originally dolomite or calcite, the nodules indicate that this was a soil of dry climate and high base status, a conclusion compatible also with the abundance of plagioclase and other easily weatherable minerals in the profile. The depth to the carbonate nodules (Bk horizon) is more like that of Alfisols than Aridisols, and this depth would have been even greater before compaction during burial. The upper part of the paleosol contains irregular ferruginized root traces up to 6 cm in diameter and burrows up to 2 cm. Its vegetation probably was dry woodland. Such vegetation may have been widespread in quartzo-feldspathic alluvium of the "Old Red Continent", an ancient land mass that included what is now Britain, western Europe, Greenland and North America.

Fossil Alfisols are common in non-marine rocks of younger geological periods (Retallack, 1986a). In many cases they also preserve fossil evidence of their ecosystems. Gleyed Alfisols may contain remains of stumps and leaf litter. Well drained fossil Alfisols, on the other hand, commonly preserve fossil bones, teeth, snails and phytoliths of plants (Retallack, 1984). Fossiliferous paleosols of these kinds show special promise for unravelling the coordinated evolution of life and landscapes (Retallack, 1988).

Spodosols

Quartzose, sandy, acidic soils may have a very distinctive subsurface spodic (Bs) horizon in which the sand grains are coated with iron and humus compounds, usually with deep radial cracking (de Coninck et al., 1974). Such characteristic horizons of Spodosols are thought to form preferentially on initially siliceous parent materials by the leaching effects of chemicals such as phenols washed from leaves by rain (Fisher and Yam, 1984). The oldest currently known Spodosols are Eocene in age (Pomerol, 1964), but the process of their formation (podzolization) was clearly initiated during Paleozoic time. Incipient podzolization can be seen in the Peas Eddy clay paleosol of Late Devonian age (Retallack, 1985). More advanced podzolization is found in the Rogerly sandy clay loam paleosol (Figures 21.5 and 21.6) of mid-Carboniferous (Namurian) age in a quarry high on the hill southwest of Stanhope, England (Percival, 1986). This profile is an Inceptisol (Dystrochrept more precisely), but is so close to qualifying as a Spodosol that discovery of Spodosols can be expected in Carboniferous rocks eventually.

The Rogerly sandy clay loam is a thick (at least 2 m) profile. Its surficial coal (O horizon) may reflect more waterlogged conditions on burial of the paleosol. There is no evidence of waterlogging during its formation. The bleached (E) horizon is

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Fig. 21.5. The type Rogerly sandy clay loam paleosol of the mid-Carboniferous (Namurian) Firestone Sill Sandstone exposed in the upper quarry east of Stanhope, England (British National Grid NZ0122383). Hammer handle in A horizon is 25 cm.

thick over an incipient spodic (Bs) horizon (Figure 21.5) and stigmarian rootlets penetrate very deeply into the subsurface. The paleosol appears to have formed on sands of the levee of a large deltaic distributary that later became a swamp and ultimately a marine bay, as do modern delta lobes of the Mississippi Delta (Coleman, 1988). Unlike the conifer forest of modern boreal Spodosols or the angiospermous dwarf forest of modern "giant podzols" in tropical regions (Richards, 1952), this paleosol supported a forest of tree lycopods whose roots (*Stigmaria ficoides*) were quite stout (up to 5 cm in diameter). Could this difference in vegetation explain the weak mobilization of iron and humus in this mid-Carboniferous Inceptisol compared to modern Spodosols? Weak development also could be due to a short time of formation, but this does not tally with the depth of leaching and abundance of root traces in this paleosol which would be unusual for a weakly developed profile.
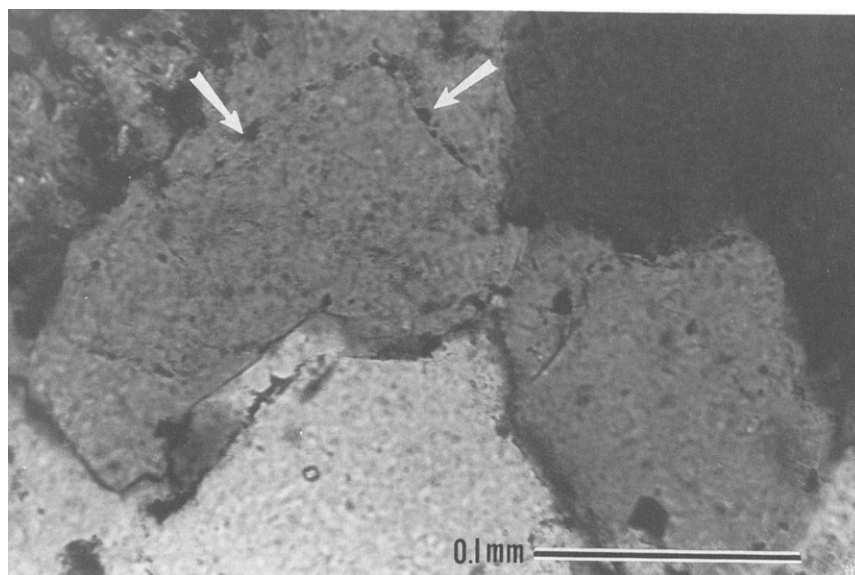
Fig. 21.6. Petrographic thin section (crossed nicols) of thin iron and humus coatings on angular grains, since overgrown with silica during burial diagenesis. From the subsurface (Bs) horizon of the type Rogerly sandy clay loam paleosol, of the mid-Carboniferous (Namurian) Firestone Sill Sandstone, near Stanhope, England. Scale bar is 0.1 mm.

A relationship between degree of expression of spodic horizons and plant evolution cannot yet be proven, but has some theoretical appeal. Phenolic substances, often implicated in podzolization, are widespread in plants as biochemical byproducts of wood production that also are effective as mild insecticides (Swain and Cooper-Driver, 1981). It is conceivable that phenolic concentrations and effectiveness have increased in conifers and angiosperms over hundreds of millions of years of evolution in the face of the ever present threat of insect attack. The fossil record of Spodosols, and related soils, may offer clues to this loosely coordinated evolution of insects, plants and soils.

Ultisols

Base-poor soils with clayey subsurface (Bt) horizons present problems of low nutrients for plants, similar to Spodosols. Ultisols also may date back to Carboniferous time (Retallack, 1986a), although none have been reported older than Triassic (Lucas, 1976) and well documented examples are no older than Eocene (Retallack, 1983). The exact timing of the afforestation of such soils is of some interest. It could be argued, as has Weaver (1967) for example, that the stabilizing influence of forests was necessary in order to promote such deeply weathered and differentiated pale-

osols. On the other hand, forests could have been established geologically early on nutrient-rich lowland soils, but only later developed microbially mediated nutrient procurement systems, such as mycorrhizae, that allowed them to grow in nutrient poor upland soils. Current information remains insufficient to decide this and related questions, but study of the trace fossils of fossil Ultisols and related soils can now be seen as a way to address such formerly intractable problems of historical geology.

Oxisols

These uniform deeply weathered soils of ancient stable land surfaces are surficial equivalents of many paleosols along major geological unconformities. Such deeply weathered soils appear to be geologically ancient. High grade metamorphic rocks at least as old as 2600 million years have the highly aluminous and ferruginous composition of Oxisols (Dash et al., 1987). The oldest little metamorphosed profiles showing such oxidation and deep weathering are some 1660 million years old in the Thelon and Athabaska Basins of north central Canada (Ramaekers, 1981; Chiarenzelli et al., 1983; Tremblay, 1983; Holland, 1984). Oxisols commonly are associated with laterite and bauxite. These also are geologically ancient. Banded iron formations have been found to have been enriched to laterite as long ago as 2070 million years in South Africa and 1800 million years in Western Australia (Morris, 1985). Highly aluminous metamorphic rocks as old as 3500 million years in the Aldan Shield region of Siberian U.S.S.R. (Salop, 1983) have been interpreted as former bauxites (Serdyuchenko, 1968). Similar explanations were advanced for corundum ores some 2700 million years old in Zimbabwe and in Western Australia, but both have been interpreted later as hydrothermally altered rocks around volcanic hot springs (Schreyer et al., 1981; Martyn and Johnston, 1986). More promising candidates for ancient bauxites are rocks rich in pyrophyllite and kyanite within the 3000 million year old Nsuze Group of the Barberton Mountain Land of South Africa (Reimer, 1986). Bauxites have been found in rocks of most geological periods up to the present (Bardossy, 1982).

When these ancient kinds of soils became colonized with plants and animals, and what effects they had on them, remain open questions. Laterites of Sri Lanka, probably dating to Late Cretaceous or Early Tertiary time have extensive channeling attributed to roots and burrowing animals (Vermaat and Bentley, 1955). Such traces need to be sought in older bauxites, laterites and Oxisols.

Both life and soils on these unusually stable cratonic landscapes have changed considerably over geological time. Although many Precambrian paleosols at unconformities are impressive in their thickness and degree of weathering, they pale against the extreme weathering and horizon differentiation extending to many tens of meters in lateritic and bauxitic profiles of Tertiary and Mesozoic age (Senior and

Mabbutt, 1979). In humid regions some Oxisols now support tropical rainforest in which nutrient difficulties of the soil are circumvented by efficient recycling in a shallow root mat (Sanford, 1987). Many deeply weathered Oxisols and laterites formed under forest now are exposed in regions where they support open grassy woodlands, such as the miombo of East Africa (Cole, 1986), or desert vegetation, such as mulga of Australia (Beadle, 1981). These are soils that form over long periods of time. Many of them have had complex histories.

Coordinated evolution of soil and life

Although studies pertinent to the evolution of soil types during Paleozoic time are few, it seems clear that this was a time of diversification of soils hand in hand with the evolution of new kinds of vegetation (Figure 21.7). Other factors could have been involved in this diversification of soil types, but probably did not have as direct an influence. Continental geography and parent materials were by and large as varied during Paleozoic time as they are today. So also was the tectonic stability of landscapes, ranging from stable ancient cratonic land surfaces to young alluvial plains and volcanic mountains (Zeigler et al., 1979). In addition, the oxygenation of the atmosphere, that left a record in the changing nature of Precambrian paleosols, appears to have been largely completed by early Paleozoic time (Feakes et al., 1989).
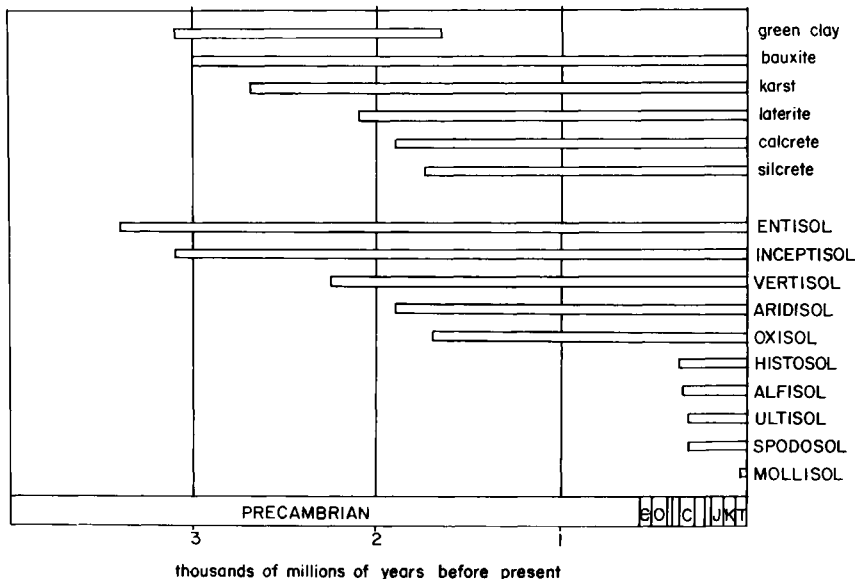

Fig. 21.7. Geological time ranges of U.S.D.A. soil orders and other weathering features. Green clay is an informal name for certain Precambrian paleosols. (Retallack, 1986a.)

These other variables in soil formation do not account for the observed diversification of Paleozoic soil types as well as the diversification of plant formations.

Paleozoic paleosols thus offer a unique view of major events in the progressive cover of the landscape by different kinds of vegetation (Retallack, 1985). Not enough is yet known to decide whether plants and animals invaded the land in successive waves from lakes and oceans (Knoll et al., 1984) or whether they evolved on land from pre-existing soil microbes (Stebbins and Hill, 1980). There are, however, hints that terrestrial ecosystems were built up by degrees: from microbial earths to polsterlands, brakelands, shrublands and woodlands. There also are indications that nutrient-rich soils, with carbonates and with easily weathered minerals such as plagioclase, were colonized earlier than siliceous sands or deeply weathered clay. Well-drained soils were occupied by polsterlands surprisingly early (late Ordovician). Marine-influenced marsh also is quite ancient (Early Silurian). Woody vegetation appeared almost simultaneously (geologically speaking, during the mid-Devonian) in marine influenced, waterlogged and well-drained soils. Afforestation of nutrient-poor peats, silica sands, kaolin clays, laterites and bauxites are less well constrained, but may have been somewhat later (latest Devonian to Triassic). Such a stepwise development of different vegetation types is understandable from ecological structure and succession of ecosystems. Microbial nutrient procurement systems, especially for phosphorus, potassium and nitrogen, probably were necessary before larger plants could thrive. A landscape stabilized by herbaceous vegetation may have been necessary before trees could grow to any stature.

The advents of these various plant formations were major biological events and were also crucial to weathering regimes. Forests, for example, create their own microclimate in which diurnal and seasonal variation in humidity and temperature are dampened to more equable levels. Forests also persist in the landscape by resisting the perturbing effects of floods, and forests also maintain global atmospheric oxidation levels. They are the most significant producers of oxygen, both as a by-product of photosynthesis and by helping to maintain and bury in swamps organic carbon that would otherwise burn or decay to carbon dioxide. When oxygen levels increase, forest fires rage. When carbon dioxide levels increase, the flammability of trees and forest fires is suppressed (Cope and Chaloner, 1985). It was during the Paleozoic that these environmental feedback systems became established. It was thus a period of geological history when there was increasing biotic control of soil formation.

Two equally extreme views on the role of life in surficial processes can be imagined, and both probably are equally incorrect. According to the Gaia hypothesis, life has always, and will continue, to exert a homeostatic influence on surficial processes that serves to promote its own persistence (Lovelock, 1979). This is an optimistic and comforting view, though somewhat unscientifically, almost mystically, holist. An equally extreme alternative suggested mainly by earth scientists (Holland, 1984) is that life is a trivial epiphenomenon that adapts as best it can through evolution to wider environmental forces, such as the differentiation of the Earth's interior and its

surface expression as volcanoes and earthquakes. This pessimistic view of life's role on Earth can be characterized as Ereban, after the primeval god of darkness (Figure 21.8; Retallack, 1990). These extreme views are not much more than a deification of an old debate in ecology (O'Neill et al., 1986). Are communities in any sense like superorganisms or are they simply random associations of species that happen to have similar environmental requirements? The answer is probably somewhere between these extremes. This can be seen from the example of forests, whose specific composition varies widely according to soil type and regional climatic history. On the other hand, forests are environmentally regulated habitats for other creatures and play a role as regulators of global atmospheric composition.

The question now before us is not whether life has absolute or no control over surficial processes, but rather how much? A good answer to this question is still a long way off, but some approaches now can be seen. One way is to evaluate the fluxes of volatile elements in modern ecosystems. Such work is showing that life does not have a monopoly on surficial processes over physical forces such as volcanoes, but life's influence is still far from negligible (Mooney et al., 1987). Another way to assess life's role is by historical evaluation of biotic influences in soil formation through geological time. As outlined here, there does appear to have been a general trend of increasing biotic influence in Paleozoic paleosols. Life does not seem to have rapidly assumed a hegemony of Earth resources soon after it evolved (an extreme Gaian view; Figure 21.8). Nor does it seem likely that life has always been at the whim of every volcanic eruption, change in sea level or major meteorite impact (an extreme Ereban view). Additional studies of Paleozoic paleosols, controlled for soil forming factors other than biotic ones, show some promise for probing the intricate interrelationship between life and soil.
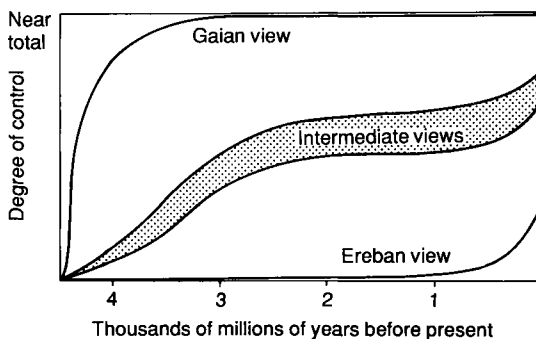

Fig. 21.8. A conceptual model for the historical degree of control by extreme Gaian and Ereban views, together with a range of more reasonable intermediate views.

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